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THE EFFECTS OF OCEANIC FRONTS ON PROPERTIES OF THE ATMOSPHERIC BOUNDARY LAYER

by

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ABSTRACT

The theories of Mosby and Åmot on the response of boundary layer properties are applied to the computation of air temperature and water vapor pressure at a height of 10 m. above the sea surface. The effects of different empirical constants on the equilibrium differences of the sea and air temperatures and water vapor pressure, were investigated numerically. The computations indicate that, for all practical purposes, the equilibrium of sea and air temperatures is reached in about five hours when the rate of change of sea surface temperature along the trajectory of the air remains constant. The numerical value of equilibrium difference depends on the rate of change of sea surface temperature along the trajectory of the air. The results of the computations are presented in graphs. The application of the Åmot-Mosby theory provides a method for synoptic numerical analysis/forecasting of surface air temperature and water vapor pressure over the oceans. The values of the empirical constants should, however, be determined more accurately from a properly devised measurement program in a variety of conditions.

In the numerical simulation of the feedback effects, the surface wind was allowed to blow across an oceanic front, both from warm to cold and from cold to warm sides. The surface air temperature, water vapor pressure of the

air, and various heat exchange components were computed. The resulting changes of the properties of surface air moving from the cold to the warm side are essentially quite different than in air moving in the opposite direction. There is a characteristic "overshooting" of air temperature and water vapor pressure at the edges of the oceanic front. The overshooting is physically and numerically explainable on the basis of the dependence of equilibrium differences (sea surface and air temperature) on the rate of change of sea surface temperature along the trajectory of the air.

The computed air temperature and heat exchange components across the oceanic front are used in the Sawyer-Bushby equation and the thermodynamic equation. The effects of feedback (at the oceanic fronts) on the state and motions in the lower layers of the atmosphere by different types of air flow were numerically simulated. These numerical experiments explain several weather and oceanographic phenomena observed at the fronts. It can be concluded that the oceanic fronts are a significant feature in the determination of the mean position of the atmospheric fronts in the proximity of oceanic polar and artic fronts.

1. PURPOSES OF THE STUDY

The exchange processes between the sea and the atmosphere are the primary cause for modification of surface layers of air over the ocean. In any energy exchange process, the differences between the properties of the surface layer of the sea and the corresponding properties of the surface layer of air determine greatly the direction and intensity of the exchange processes.

Thus, the first purpose of this paper is to study numerically the rate of modification of the properties of surface air moving over the ocean and to establish and test numerical methods for synoptic computation of surface air temperature and water vapor pressure.

The horizontal gradients of heating and cooling of surface air masses create pressure gradients which affect the movement of air. These horizontal gradients of heating and cooling are expected to be most pronounced where the underlying surface has sharp thermal gradients (i.e. at oceanic fronts). Surface winds in excess of twice the gradient wind are often observed at oceanic frontal regions (Cormier and Kindle, 1971). It is of further interest to point out that the mean position of atmospheric polar and arctic fronts over the oceans is in the vicinity of the oceanic polar and arctic fronts. Consequently, the second objective of this paper is to report on a preliminary numerical study of the

effects of oceanic (polar) fronts on the atmosphere and to study the effects of exchange processes on other frontal phenomena such as mesoscale local winds and fogs.

Use will be made only of properties which are measured during routine (synoptic) weather observations or which can be easily derived from these observations so that the results can be directly applied to routine weather and oceanographic analyses/forecasts.

2. THEORY OF THE LARGE-SCALE RESPONSE OF SURFACE AIR TEMPERATURE AND WATER VAPOR PRESSURE TO THE CORRESPONDING PROPERTIES OF THE SEA SURFACE

The various theories of the response of maritime air to the properties of the sea surface were reviewed by Roll (1965). A Norwegian theory originally reported by Åmot (1944) and later by Mosby (1957) and Bøyum (1962) is presented here with a few modifications. This theory is adaptable to numerical computations using parameters available in routine meteorological observations. In this theory the trajectory of the surface air is followed. The surface air properties generally refer to a standard height of 10 m. above the sea surface. The detailed profile of these properties with height (lapse rate) at any given time is not specified as it is not available from routine synoptic observations. The theory can be considered as one of turbulent transfer in which the intensity of turbulent diffusion is related to the "transfer potential" (the difference between the properties of the sea surface and those of the air) and wind speed.

An air parcel moving over the sea surface has its temperature changed mainly by the convective turbulent transfer of sensible heat. If the temperature of the sea surface (T_w) changes along the trajectory of the air, the corresponding individual change of sea-air temperature difference $\Delta T = T_w - T_a$ is:

$$\frac{d\Delta T}{dt} = \frac{dTw}{dt} - \frac{dTa}{dt} \qquad (1)$$

where T_a is the air temperature.

The change due to sea-surface temperature effects along the trajectory is essentially composed of two terms:

$$\frac{dTw}{dt} = \frac{\partial Tw}{\partial t} + \mathbb{V} \cdot \nabla T_w \qquad (2)$$

The "local change" of sea surface temperature $\frac{\partial Ta}{\partial t}$ is small (usually much smaller than 0.01° C./hour) and can be neglected especially in comparison with the last term ($\mathbb{V} \cdot \nabla T_w$), which represents the sea surface temperature change along the trajectory of the surface air. $\mathbb{V}$ is the velocity of the surface air parcel and can be assumed to be the surface wind speed. Thus equation (1) reduces to:

$$\frac{d\Delta T}{dt} = \mathbb{V} \cdot \nabla T_w - \frac{dT_a}{dt} \qquad (3)$$

Synoptic sea surface temperature analyses as well as analyses of surface wind are available routinely.

The individual change of air temperature $\frac{dT_a}{dt}$ can be obtained from the first law of thermodynamics:

$$\frac{dT_a}{dt} = \frac{dQ}{c_p dt} + \frac{R\ T_a}{c_p\ p}\frac{dp}{dt} \qquad (4)$$

where R is the gas constant, p the pressure, c_p the specific heat of the air, and Q the addition of heat. The last term

includes adiabatic cooling of air by advection, the magnitude of which can be computed for various conditions. This computation has been performed by Åmot who found that advective adiabatic cooling of the surface air is insignificant for moderate wind velocities. At 55° N. Åmot found that if $\frac{\partial p}{\partial t}$ is as large as 1 mb. per hour, then

$$\frac{RT_a \partial p}{C_p p \partial t} = 0.078° \text{ C per hour.}$$

In the present sea-air interaction problem the direct radiative heating and cooling of the air will be neglected.

The convective transfer of sensible heat is a function of ΔT (see e.g. Laevastu, 1960). Furthermore, Mosby (1933) and Bøyum (1962) found from observations that the changes of air temperature can be well represented as:

$$\frac{dT_a}{dt} = K (T_w - T_a) - C \qquad (5)$$

where K and C are positive constants. Substituting (5) into (3), re-arranging, and assuming $\mathbb{V} \cdot \nabla T_w$ to be constant, equation (3) can be transformed into the integrable form:

$$\frac{d\Delta T}{dt} + K \Delta T = C + \mathbb{V} \cdot \nabla T_w \qquad (6)$$

which after integration gives:

$$\Delta T = \Delta T_o e^{-Kt} + \left(\frac{C}{K} + \frac{1}{K} V_r \frac{\partial T_w}{\partial r} \right) (1-e^{-Kt}) \tag{7}$$

when r is distance. This is a useful formula for computation of ΔT where ΔT_o is at the time T_o. The empirical constants of C and K must be determined from observational data.

We can also derive a useful formulation for numerical computation of surface air temperature over the ocean if we assume as before that the temperature of an individual parcel of air is changed proportionally to the sea-air temperature difference (an assumption analogous to Mosby's (1933) empirical results):

$$\frac{dT_a}{dt} = K (T_w - T_a) - C \tag{5a}$$

Expanding the term on the left, the equation becomes:

$$\frac{\partial T_a}{\partial t} + \mathbb{V} \cdot \Delta T_a = K (T_w - T_a) - C \tag{8}$$

Equations (8) and (5) are equivalent. Equation (6) and (7) are frequently used for research purposes and equation (8) is normally applied operationally.

Equation (8) can be expanded to another form. Assume that the local sea surface temperature does not change with time: $\frac{\partial T_w}{\partial t} = 0$ and add this zero term to the left side.

Add the term $\mathbb{V} \cdot \nabla T_w$ to both sides of the equation:

$$\frac{\partial T_w}{\partial t} - \frac{\partial T_a}{\partial t} + \mathbb{V} \cdot \nabla T_w - \mathbb{V} \cdot \nabla T_a = -K(T_w - T_a) + C$$

$$+ \mathbb{V} \cdot \nabla T_w \qquad (9)$$

$$\frac{(T_w - T_a)}{\partial t} + \mathbb{V} \cdot \nabla (T_w - T_a) + K\ (T_w - T_a) - C$$

$$= \mathbb{V} \cdot \nabla T_w \qquad (10)$$

$$\frac{d(T_w - T_a)}{dt} + K(T_w - T_a) - C = \mathbb{V} \cdot \nabla T_w \qquad (11)$$

The same argument as used for air temperature change can be used for water vapor pressure (e_a) change over water. The theory might be somewhat more valid for e_a as there are no appreciable radiation effects on e_a. The empirical constants (C and K) must, however, be different for e_a. The values of the constants reported in the referenced literature are summarized in Table 1. Obviously, additional investigations are desirable for determination of accurate values for these constants.

Other theories and approaches for prediction of air temperature changes over the water were also evaluated. Considering the heat content of a column of air moving across the sea surface, Burke (1945) derived an integral formula (which he evaluated graphically in a series of charts) to predict the temperature of the air as a function

Table 1

Values of Constants K and C in Equation (5) as Reported in the Referenced Literature.

Parameter	K	C	Remarks (including height of measurements)	Author
T_a	0.28	0.13	6 m. height	Mosby, 1933
	0.11	0.07	24 m. height	Mosby, 1942
	0.21	0.15	10 m. height	Bøyum, 1962
	0.064	0.095	Assumed 10 m. Obtained from monthly mean analysis of T_a and e_a.	Laevastu, 1965
	0.12	0.10	10 m. (derived by "tuning" of synoptic analysis	Carstensen and Laevastu, 1966
e_a	0.16	0.30		Bøyum, 1962
	0.15	0.18	10 m. (derived by "tuning" of synoptic analysis)	Carstensen and Laevastu, 1966

of initial air temperature, length of trajectory over the sea, initial lapse rate, and average sea surface temperature. He then used the charts to predict air temperature in the transformation of polar continental air to polar maritime air. Although the predictions close to the eastern coasts of the continents (where large changes are taking place) were relatively good, his charts were rather difficult to use over the major parts of the open oceans -- especially at lower latitudes. The charts also underestimated the final air temperature in these areas. Furthermore the formulas are not well suited for operational computer work, primarily because of difficulties in obtaining the correct initial lapse rate near the surface and in determining a realistic trajectory of the air.

Another formula for estimating the air temperature from the surface temperature has been proposed by Duquet (1960). The formula, although correct and elegant, is quite cumbersome to use. Its simplified form requires the use of a lapse rate so that his formula leads to the same difficulties as encountered by Burke.

3. THE SPEED OF RESPONSE TO T_a AND e_a TO T_w AND e_w AND THE AIR-SEA EQUILIBRIUM VALUE

Equation (7) is used for computation of the change of ΔT with time when air is moving over the water using various combinations of C and K and various rates of sea surface temperature change ($V_r \frac{\partial T_w}{\partial r}$) designated hereafter as ΔT_w. Some of the results are shown in Figures 1 through 4.

The difference $\Delta T = T_w - T_a$ (and corresponding $e_w - e_a$) reaches an "equilibrium value" (for most practical purposes) after about 5 hours and this equilibrium value is independent of the initial ΔT, but dependent on ΔT_w and on the constants C and K (compare Figures 1 and 3). The equilibrium values with different ΔT_w and different constants of C and K are shown in Figure 4. The observed equilibrium values of ΔT and Δe at Weather Ship M (from Bøyum, 1962) with different ΔT_w and Δe_w (i.e. in this case with different wind directions and speeds) are shown for comparison in Figures 5 and 6 respectively.

As there is still some uncertainty about the most accurate C and K values to be used in synoptic numerical computations, it was found desirable to demonstrate the effects of changing C with constant K as well as changing K with constant values of C. The results of this computation are given in Figure 7.

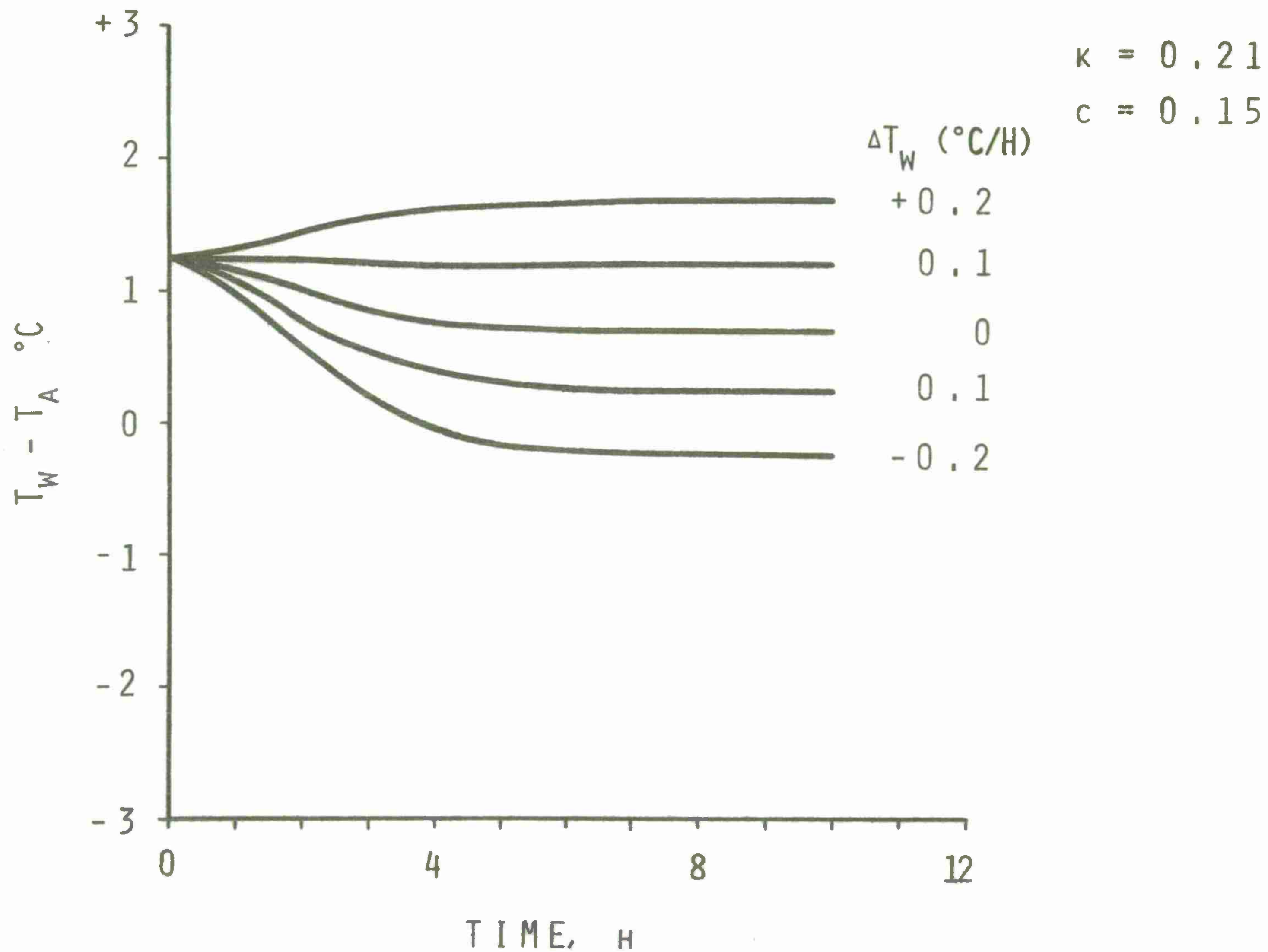

Figure 1. Change of ΔT With Time and With Different ΔT_w; C=0.15, K=0.21, Initial ΔT=1.25° C.

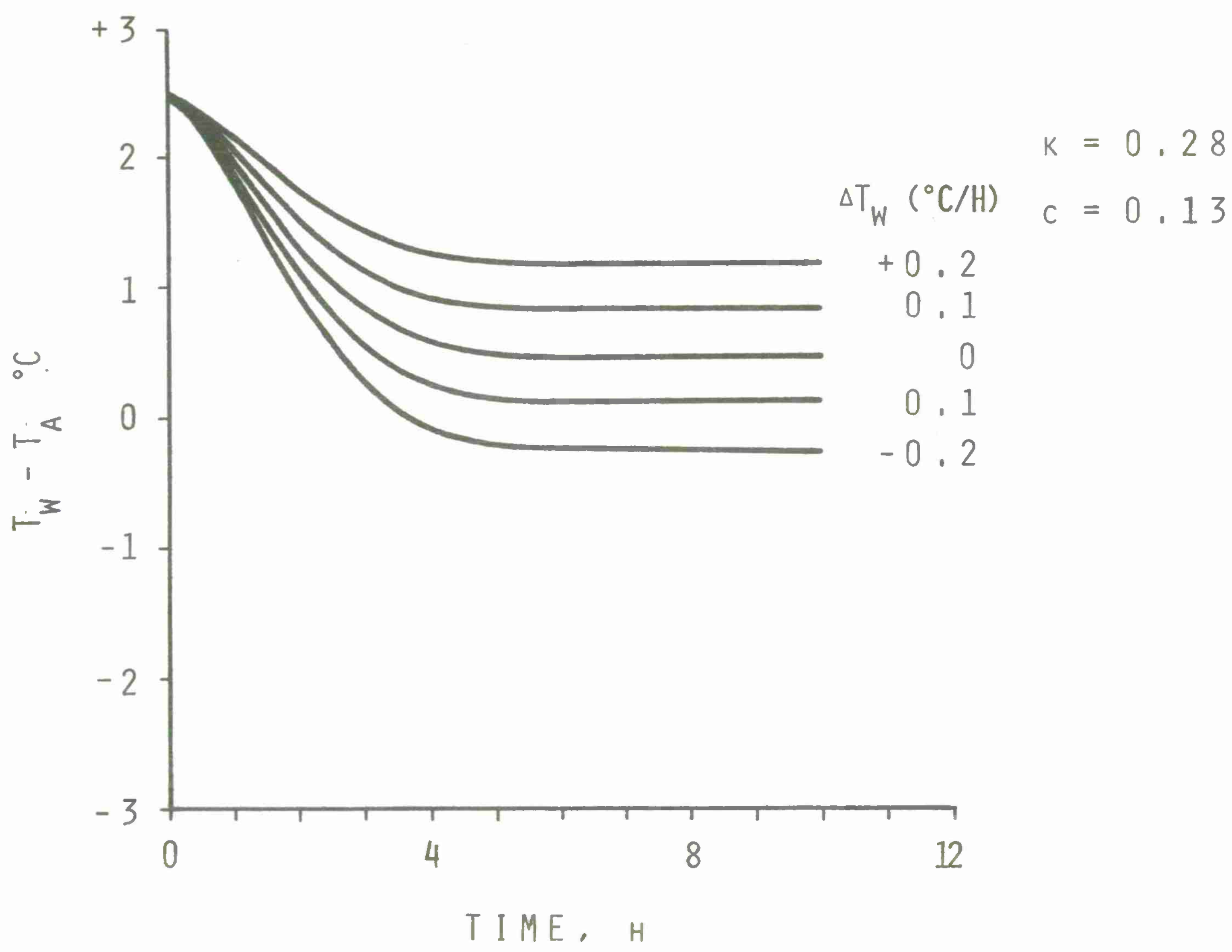

Figure 2. Change of ΔT With Time and With Different ΔT_W; C=0.13, K=0.28, Initial ΔT=2.5° C.

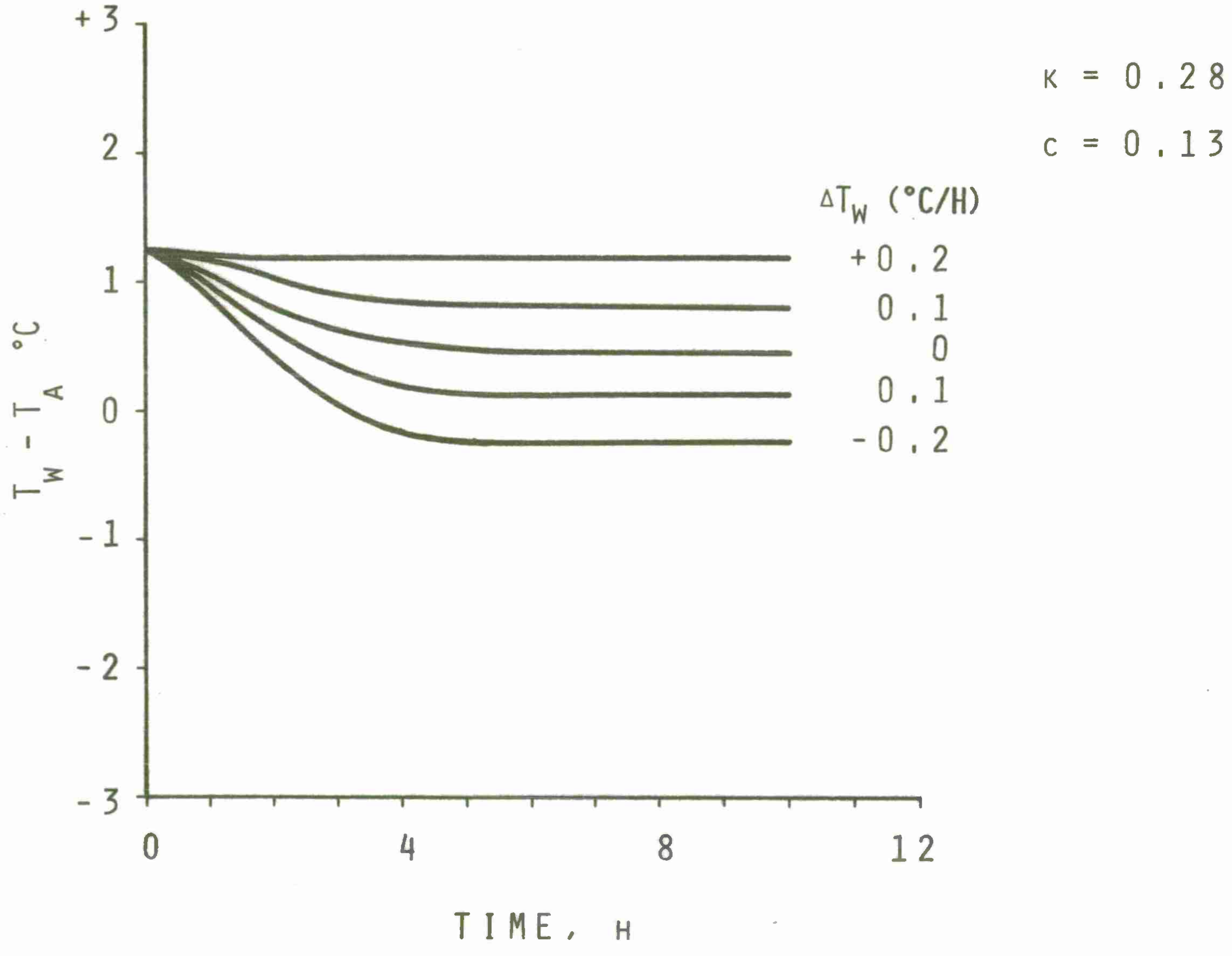

Figure 3. Change of ΔT With Time and With Different ΔT_W; C=0.13, K=0.28, Initial ΔT=1.25° C.

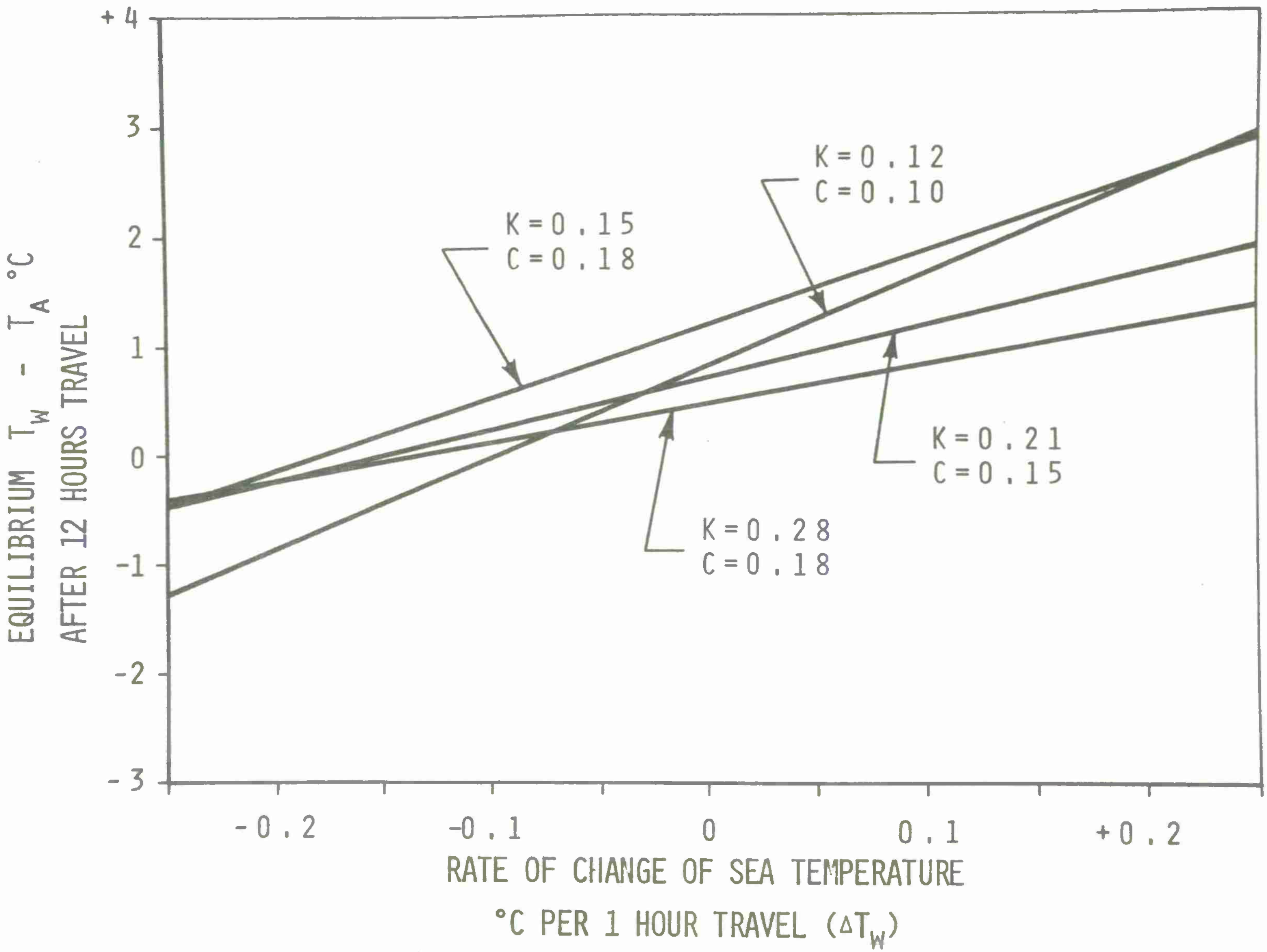

Figure 4. Comparison of Equilibrium ΔT With Different Constants of C and K and With Different ΔT_W.

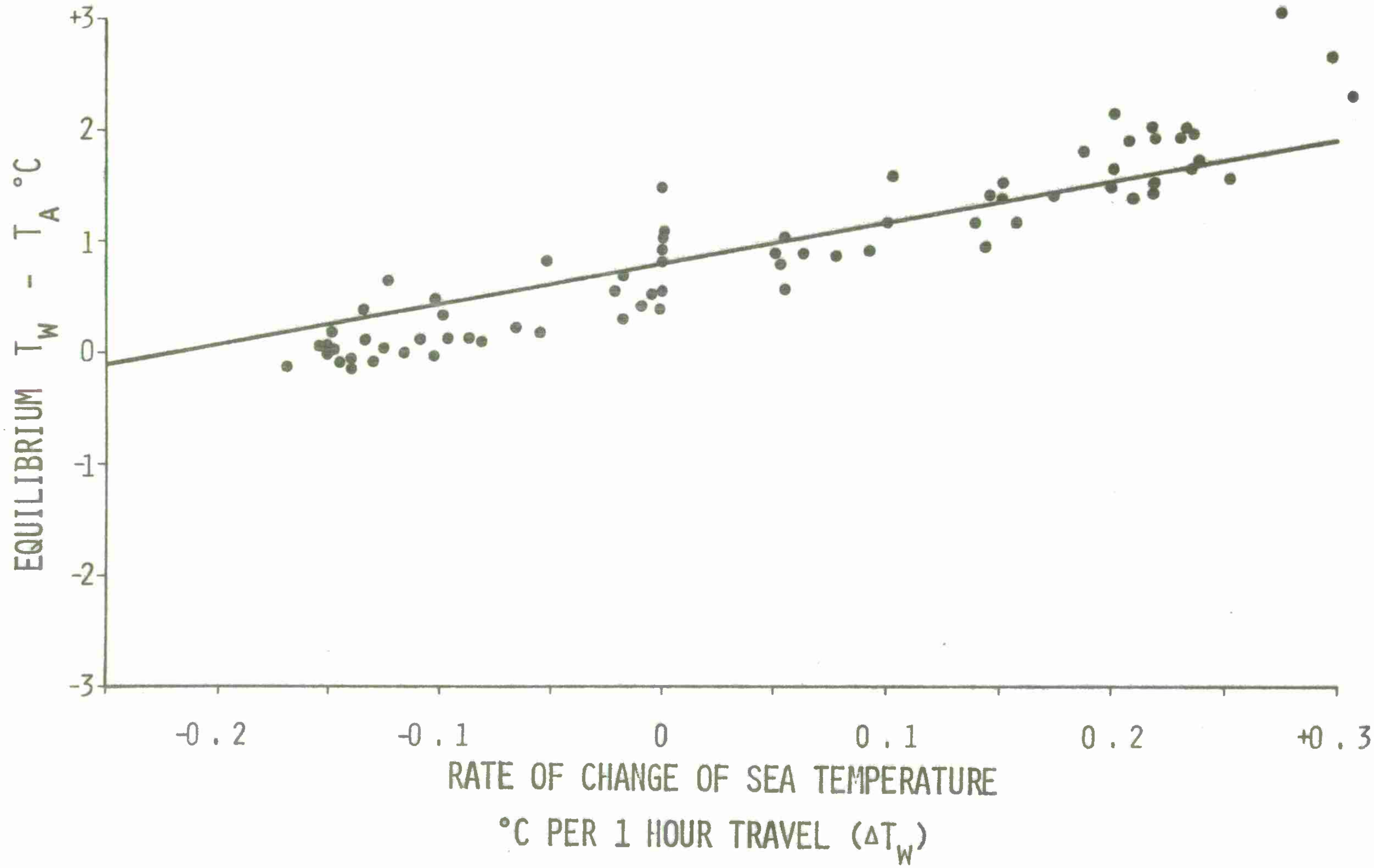

Figure 5. Equilibrium Values of ΔT With Different ΔT_W Observed by Bøyum (1962).

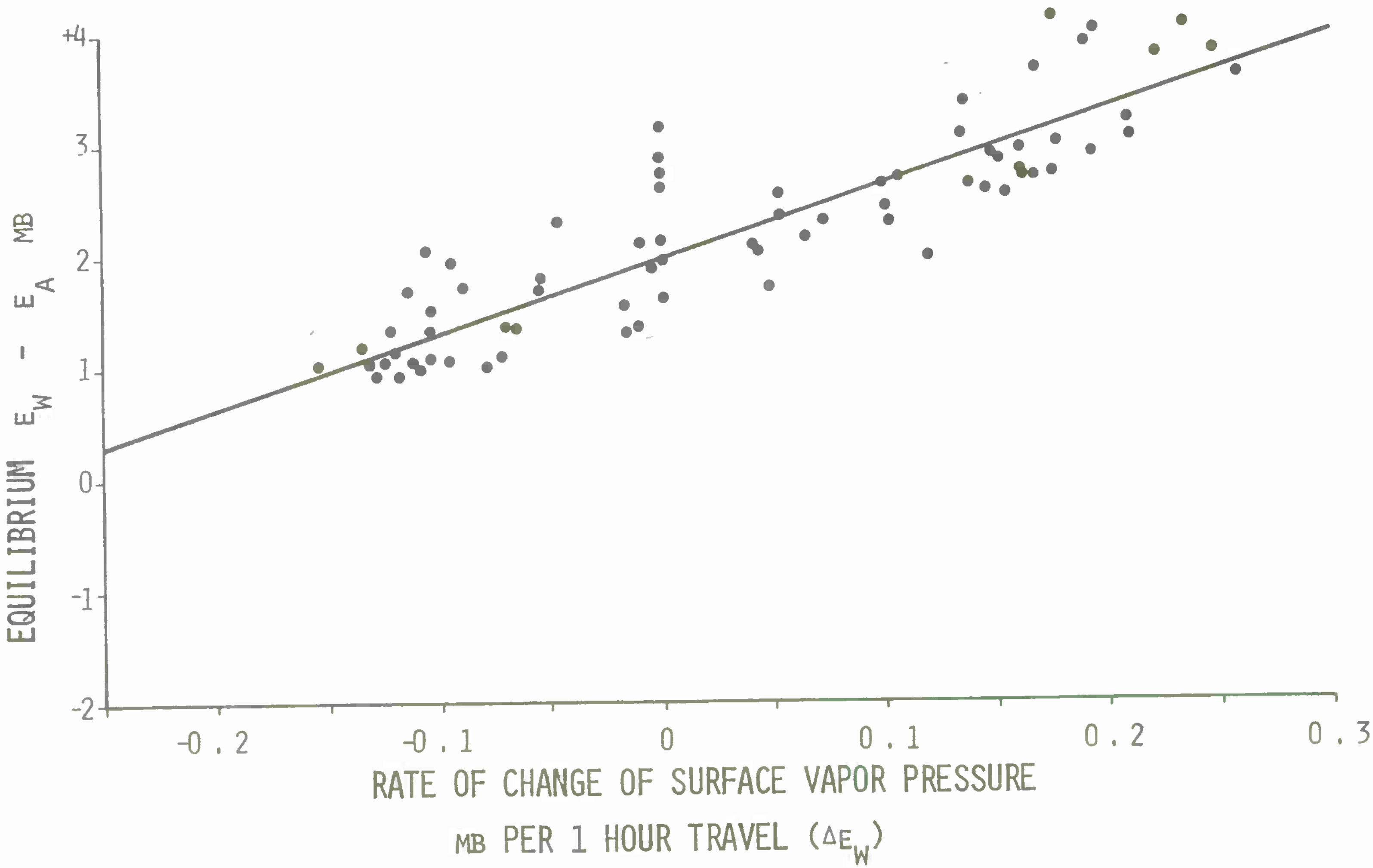

Figure 6. Equilibrium Values of Δe With Different Δe_w Observed by Bøyum (1962).

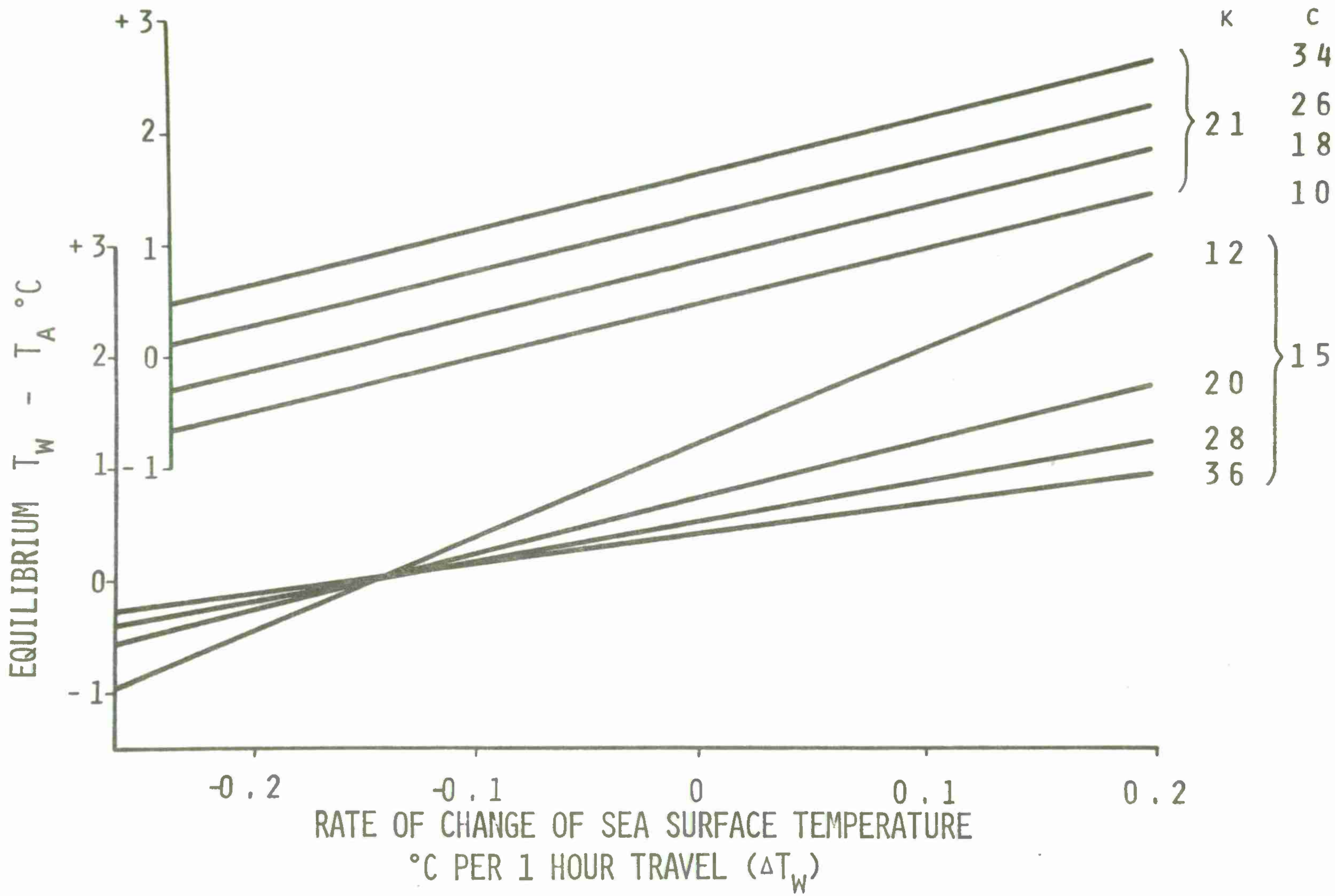

Figure 7. The Effect of Changing C With Constant K on the Equilibrium Values and the Effect of Changing K With Constant C on the Equilibrium Values.

Figures 1 through 3 indicate that when ΔT (i.e. $T_w - T_a$) reaches an equilibrium value, the turbulent transfer coefficient, which is proportional to ΔT, becomes a function of ΔT_w. This is evident in the increase of equilibrium ΔT with increasing values of ΔT_w.

It is of interest to point out that when air moves towards colder water (negative ΔT_w), the difference between the sea and air temperature (ΔT) decreases. This has been observed in nature and can be explained physically by the increase of stability of the lower layers of the air and consequent supression of turbulent transfer caused by cooling of the air close to the surface. The evaporation rate is suppressed in the same manner. There must also be a time lag in the adjustment of the properties of the surface air to the properties of the sea surface.

When the air moves towards warmer water (positive ΔT_w) then ΔT increases. This is physically explainable by the increase of turbulence caused by heating from below.

Numerical synoptic analysis/forecasting of T_a and e_a has been carried out at Fleet Numerical Weather Central, Monterey from 1965 to 1969 using the method of equation (8) (see also Carstensen and Laevastu, 1966). Some verifications of these analyses are shown in Figures 8 and 9.

Åmot (1944) presented the argument that the heat exchange between the sea and the air should be zero at the equilibrium value of ΔT when ΔT_w equals 0 (i.e. the air

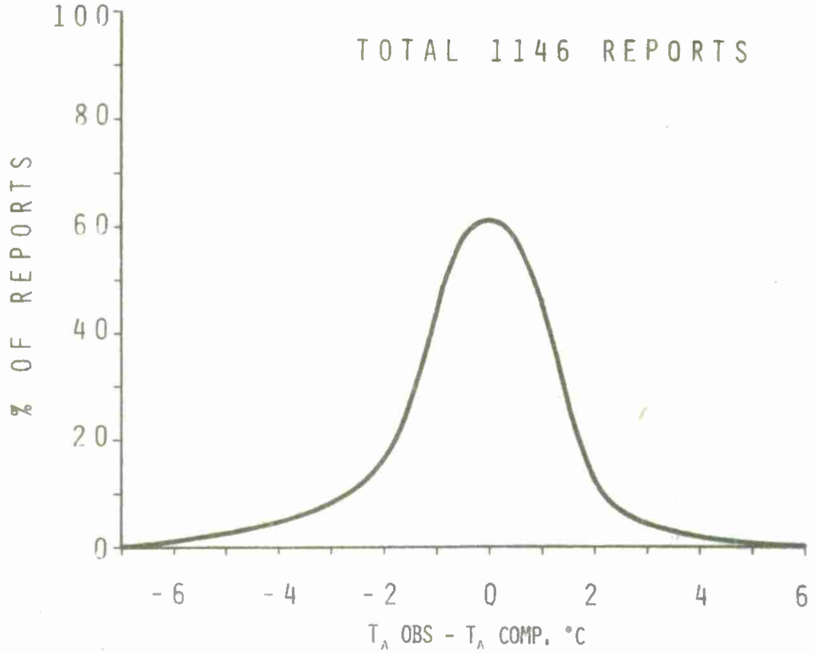

Figure 8. Difference Between Reported Air Temperature Minus Analyzed Air Temperature on 15Z, 18Z and and 21Z, 28 March and 00Z 29 March 1966.

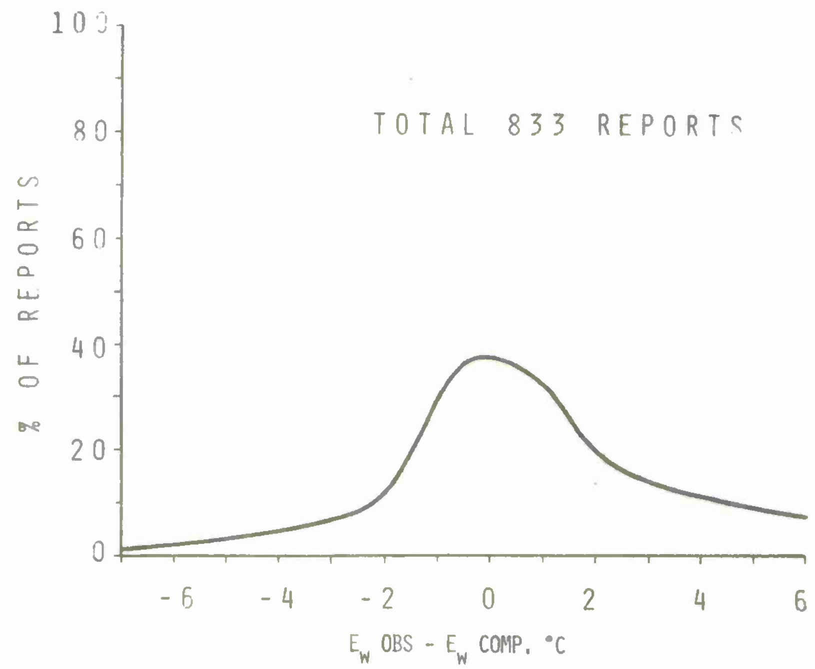

Figure 9. Difference Between Reported Water Vapor Pressure of the Air Minus Analyzed Water Vapor Pressure on 15Z, 18Z and 21Z, 28 March and 00Z 29 March 1966.

moves along sea surface isotherms). Using Mosby's measurements (1933 and 1942) at two different heights above the sea (6 and 24 m.), Åmot found that the difference in temperature was 0.01° C. per meter, which equals the dry adiabitic lapse rate. The heat exchange is assumed to be zero under dry adiabatic conditions if the vertical transport of heat by direct conduction and radiation is small compared to turbulent transport. If t increases, then equation (7) tends to become $\Delta T = \frac{C}{K} + \frac{1}{K} V_r \frac{\partial T_w}{\partial r}$. If air moves along the isotherms $\frac{\partial T_w}{\partial r} = 0$ or it is calm ($V_r = 0$), then the term containing these values becomes 0 and $\Delta T = \frac{C}{K}$ in equilibrium conditions. In these circumstances, equation (5) takes the form $\frac{dT_a}{dt} = 0$ and no heating or cooling of the lower layer of the air occurs.

Mosby measured ΔT data at heights of 6 and 24 meters and found that $\Delta T_6 = 0.46 + 3.6 V_r \frac{\partial T_w}{\partial r}$ and $\Delta T_{24} = 0.64 + 8.9 V_r \frac{\partial T_w}{\partial r}$ respectively. Assuming $V_r \frac{\partial T_w}{\partial r}$ to be 0 under equilibrium conditions as mentioned above, Åmot concluded from Mosby's measurements that heat exchange between the sea and atmosphere should be zero when $\Delta T = \frac{C}{K} = 0.40 + 0.01h$, where h is in meters. This assumption postulates that $\frac{C}{K}$ changes linearily with height.

Bøyum reasoned that the heat exchange may be zero and equilibrium conditions may prevail even when a temperature difference in excess of the adiabatic rate exists between

the sea surface and the air. The sea surface temperature is normally measured by bucket or intake thermometers and these methods do not measure the real temperature of the cooler surface film. A cooler surface film is known to exist in laboratory conditions and is caused by evaporation and sensible heat loss from the surface film. Furthermore, the sea surface temperatures measured with intake thermometers are known to have a positive bias.

The problem of the cooler surface film has not been fully explored in natural conditions and the film temperature in various wind and sea surface roughness is not known with any reasonable accuracy.

It should also be emphasized that an equilibrium condition does not necessarily mean that no exchange takes place between the sea and the atmosphere. The loss of heat from the sea can equal the removal of heat from the atmosphere.

4. EFFECTS OF AIR FLOW ACROSS OCEANIC FRONTS

In a numerical study (to be published) of the effects of sea surface temperature anomalies on the feedback of energy from the sea to the atmosphere, it became obvious that the greatest effects on the atmosphere must be found at oceanic fronts. (For a summary on the oceanic fronts see Laevastu and La Fond, 1970.)

In the numerical simulation of the oceanic frontal effects using equation (7), an oceanic front of 4° C. breadth (16° to 20° C.), a 0.5° C./10 km. gradient, and a 10 km. computational grid size were selected. A surface wind of 5.55 m./sec. was programmed to blow from the cold to the warm side in one experiment and vice versa in a second. The saturation water vapor pressure of the sea surface was computed from the prescribed sea surface temperature (T_w) and the surface air temperature (T_a). Water vapor pressure (e_a) was computed using equation (7) and coefficients of C = .15, K = .21 for T_a and C = .18, K = 15 for e_a. Some of the results are given in Figures 10 through 13.

When the wind direction was from the warm toward the colder side (Figures 10 and 11) the differences of $\Delta T = (T_w - T_a)$ and $\Delta e = (e_w - e_a)$ decreased at the southern edge of the oceanic front and increased at the northern edge. Thus an apparent "overshooting" of Δt and Δe occur at the edges of the front.

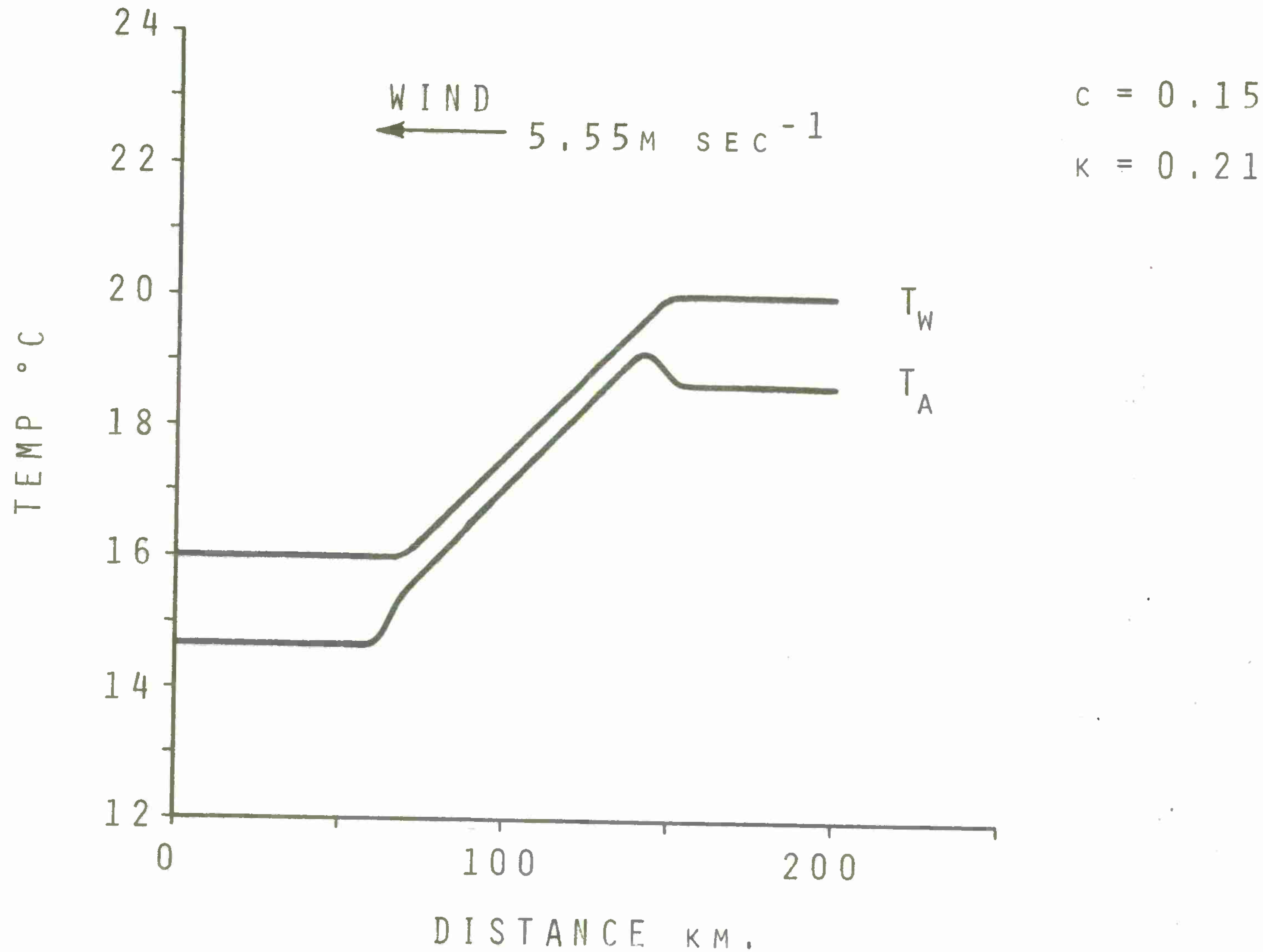

Figure 10. Change of T_a Across a Front After Wind has been Blowing From Warm to Cold Side 1 Hour.

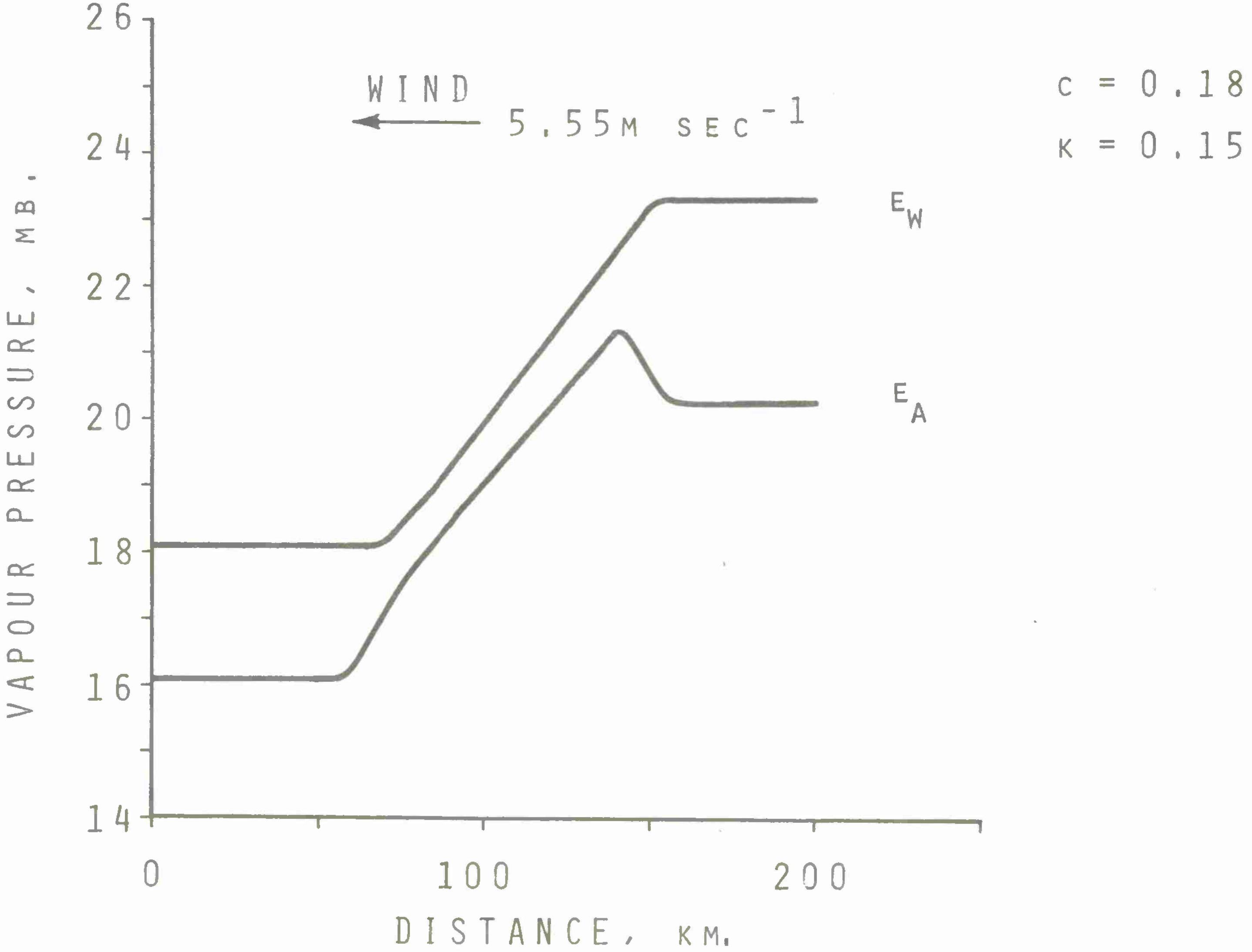

Figure 11. Change of e_a Across a Front After Wind has been Blowing From Warm to Cold Side 1 Hour.

When the air moves from the cold to the warmer side (Figures 12 and 13) ΔT and Δe increase at the north side of the front and decrease at the southern edge.

"Overshooting" at the edge of the sea surface temperature gradient is actually an adjustment process of equilibrium, ΔT to various rates of changes of ΔT_w along the trajectory (see Figures 1 through 3). As the air on the warm side in the area of weak temperature gradient moves across the front, ΔT_w becomes negative and ΔT decreases. At the cold side, ΔT_w becomes weaker and ΔT increases. The "overshooting" in air moving from the cold to the warm side can be similarly explained. Several consequences of this "overshooting" result, some of which are explored in Section 5.

The response of the surface air to the properties of the sea surface with respect to the direction of air flow across the oceanic front cause differential heating and cooling of the air and affect the heat exchange. The computation of sensible and latent heat was accomplished with the following formulae (Laevastu 1960):

$$Q_h = 39(0.26 + 0.077V)\ (T_w - T_a) \tag{13}$$

$$Q_e = (0.26 + 0.077V)\ (e_w - e_a)\ L_t \tag{14}$$

where Q_h is sensible heat exchange (g cal. cm.$^{-2}$ (24 h)$^{-1}$), Q_e is latent heat exchange, V is wind speed (m. sec.$^{-1}$), and L_t is the heat of vaporization.

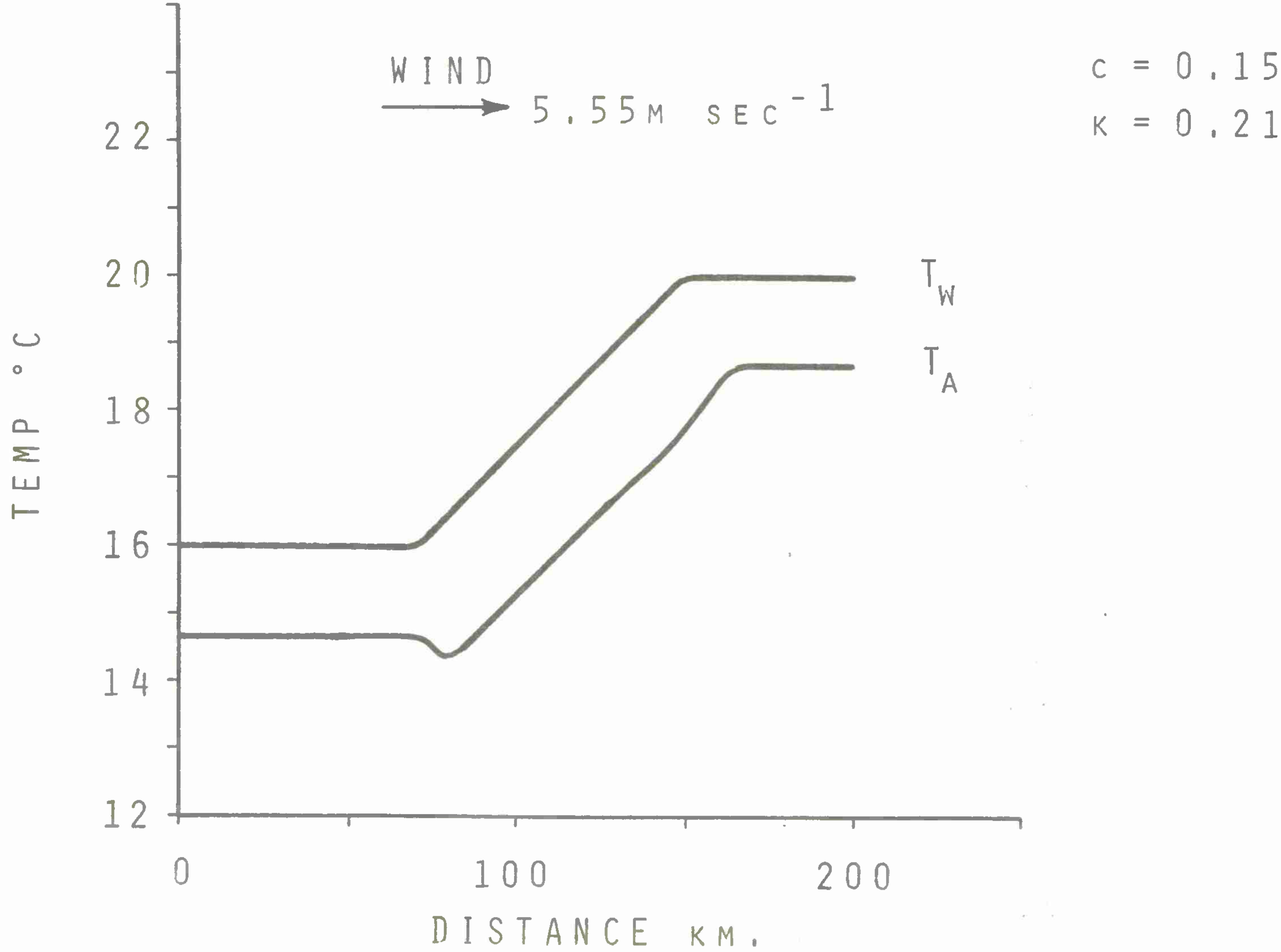

Figure 12. Change of T_a Across a Front After Wind has been Blowing From Cold to Warm Side 1 Hour.

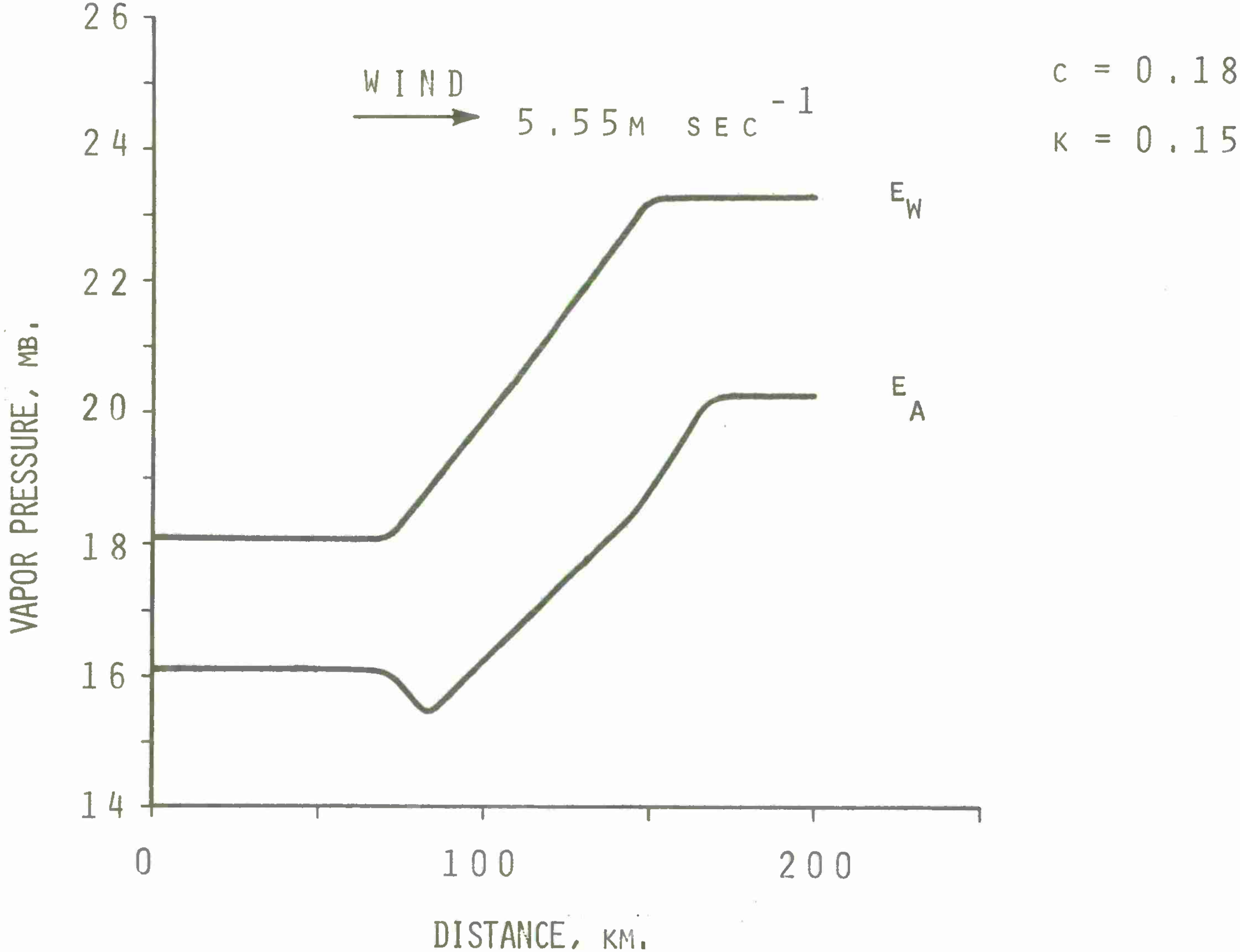

Figure 13. Change of e_a Across a Front After Wind has been Blowing From Cold to Warm Side 1 Hour.

The results of the computations are given in Figures 14 and 15. When the wind blows from the warm to the cold side, ΔT and Δe are decreased (Figures 10 and 11) and the corresponding heat exchange components are reduced (Figure 14). The opposite is the case when wind blows from the cold to the warm side (Figure 15). The latter process provides additional moisture to the warm side of the front through advection and the increase in Q_e which results in upward motions and cloud formation.

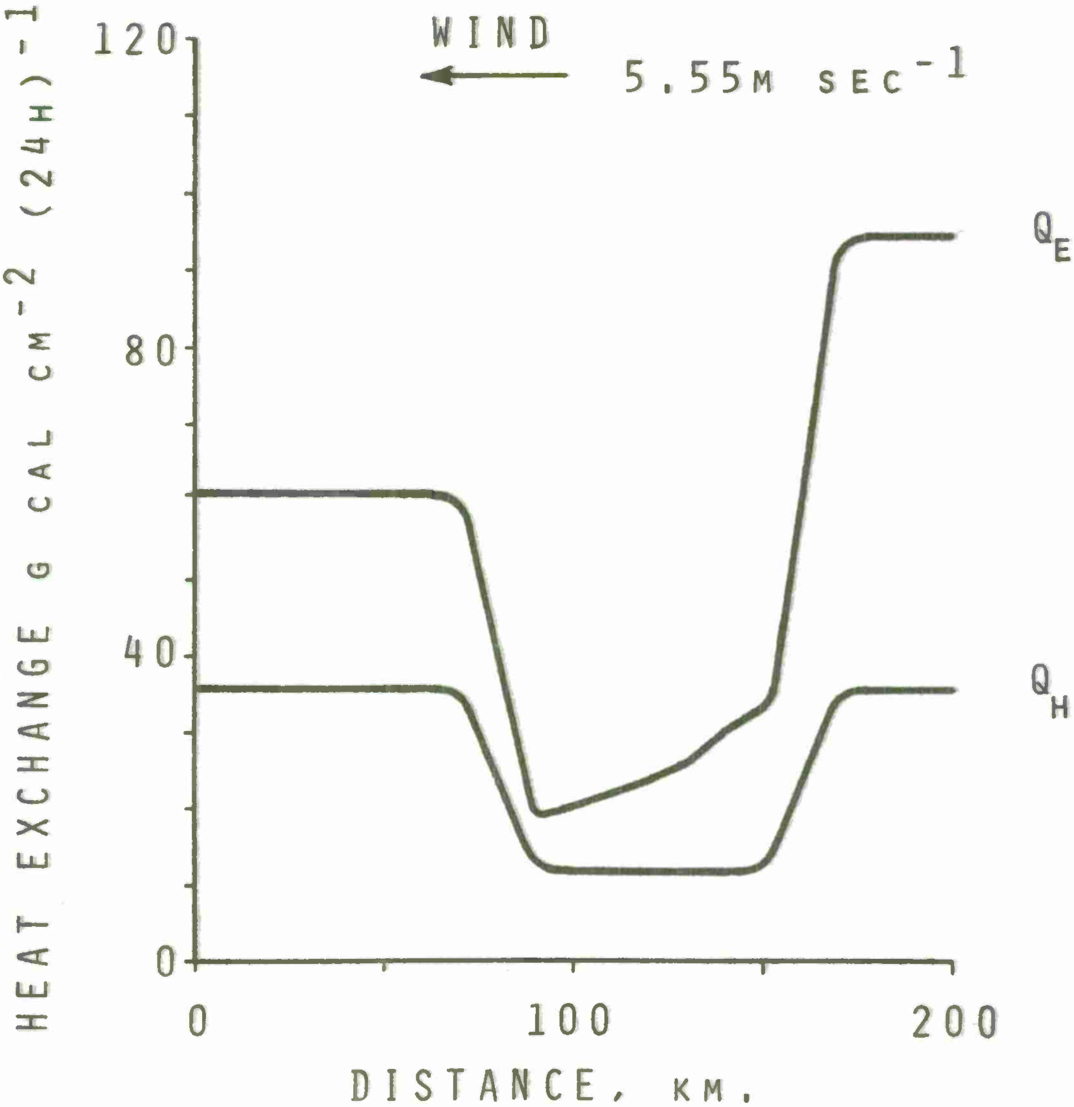

Figure 14. Sensible and Latent Heat Exchange Across a Front; Wind Blowing From Warm to Cold Side.

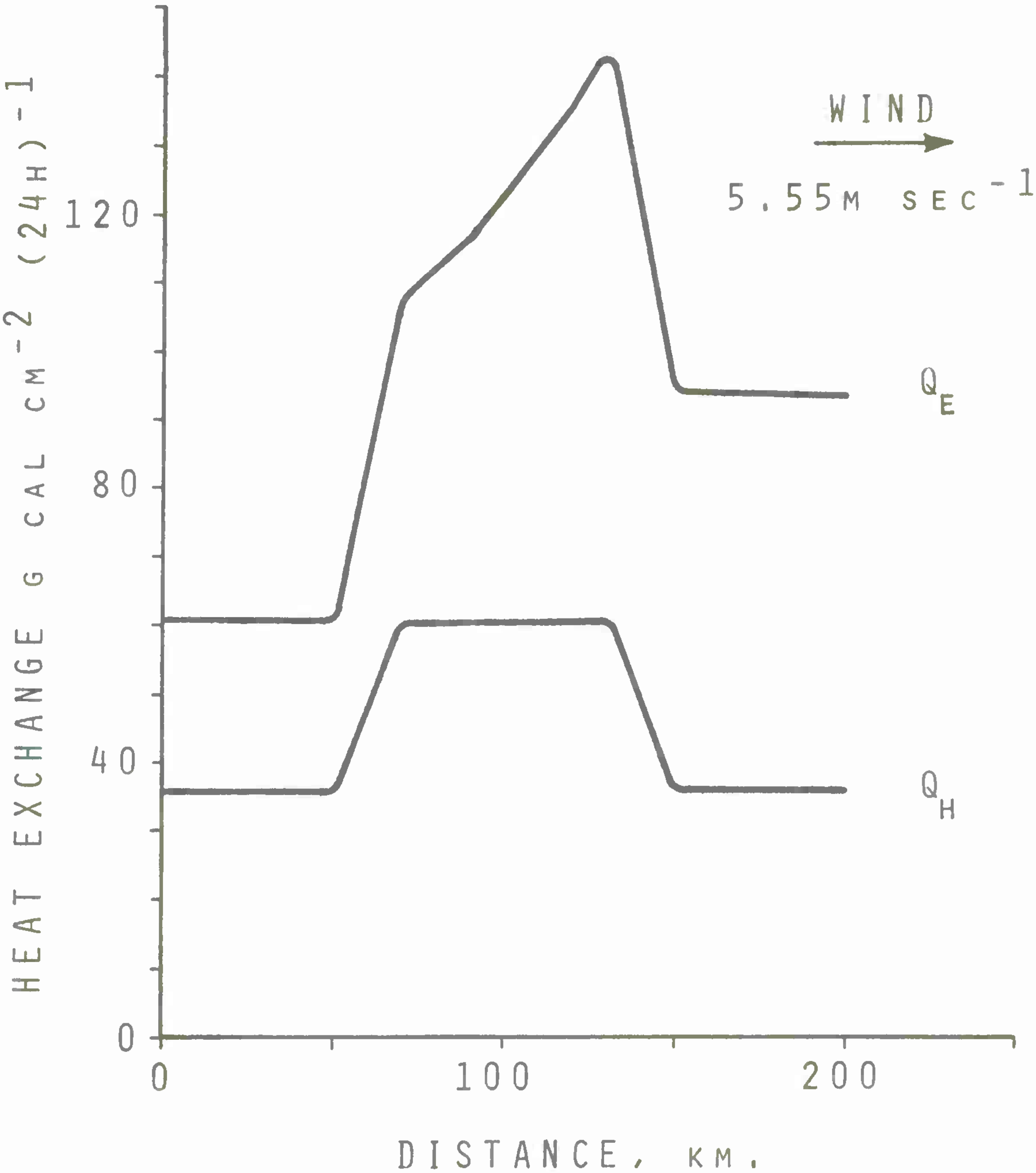

Figure 15. Sensible and Latent Heat Exchange Across a Front; Wind Blowing From Cold to Warm Side.

5. NUMERICAL SIMULATION OF SURFACE PRESSURE CHANGE, VERTICAL MOTION, AND "ZUSATZWIND" AT OCEANIC FRONTS

The numerical simulation of the feedback effects, briefly reported below, was aimed at obtaining a simplified semiquantitative picture of the nature of these effects at oceanic frontal regions.

The equation of state

$$P = \frac{\rho R' T^*}{T} \tag{15}$$

where $T^* = \dfrac{T}{(1 - 0.379\ e_{a/p})}$ indicates that pressure changes across an oceanic front can be expected due to the effect of the fronts on the differential heating and/or cooling of the lower layers of the air.

The first law of thermodynamics (equation (4)) in the adiabatic form:

$$dQ = C_p dT - \frac{1}{p}\, dp \tag{16}$$

indicates about a 2 mb. lowering of pressure during the travel of air from the warm to the cold side, across the front. Cormier and Kindle (1971), using different approaches found that simulated effects in their numerical models were less than the observed effects.

The Sawyer and Bushby (see Sawyer, 1963) equation was used for approximations of the vertical velocity (ω_Q) caused by surface heating:

$$\omega_Q \approx \frac{g^2 h^1}{Tc_p \; (P_0 - P_1) \; RA\Gamma\rho} Q \qquad (17)$$

where:

h^1 - thickness between pressure levels P_0 and P_1

$\Gamma\rho$ - departure of the lapse rate from the adiabatic

A - constant, depending on P_0 and P_1

R - gas constant

C_p - specific heat

g - acceleration of gravity

Q - heat

T - mean temperature of the column.

The results of the computations are shown in Figure 16 and indicate upward motion when the wind blows toward the warm side and downward motion when the wind is in the opposite direction. As pointed by Sawyer (1965) it is at present not clear how baroclinic disturbances caused by heat exchange at the surface affect the vertically integrated flow. The column of air is affected not only by convergence at one level but also by the balancing divergence at another level. However, increased upward motion in an area (such as caused by air flowing from the cold towards the warm side of an oceanic front) will result in an increased generation of cyclonic vorticity near the ground. This may explain the well-observed fact that extratropical cyclones tend to intensify and accelerate near oceanic frontal regions.

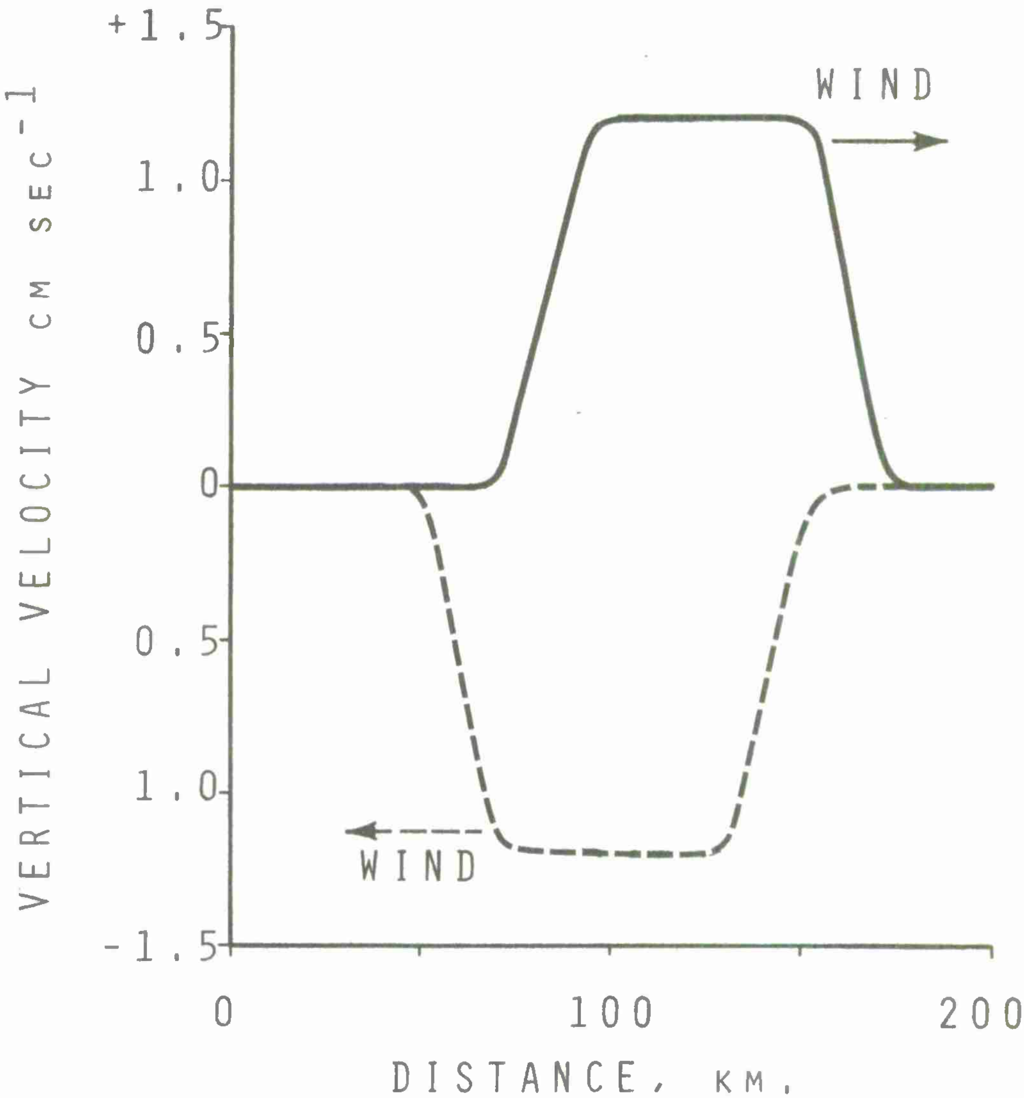

Figure 16. Estimate of Vertical Motion at the Front, Using Sawyer-Bushby Equation.

The simplified computation of pressure change across an oceanic front cannot be considered fully applicable to real nature but will, however, give indications of processes in frontal regions. The change in pressure gradient across the front caused by heating or cooling of the surface layer or by vertical motion will cause an additional wind component referred to by the German term "zusatzwind". This wind component was computed numerically using the geostrophic wind equation and the computed pressure change which resulted from heating or cooling of the lower layer of the air. In the case of the Sawyer, Bushby equation, the vertical motion was first converted to pressure change. An example of the zusatzwind computed from the equation of state is given in Figure 17. Areas of convergence and divergence between the wind and the zusatzwind are shown. If the wind direction is reversed, the area of acceleration and convergence remain in the same relative positions to the basic wind direction, but change in relation to the orientation of the front (i.e., the convergence moves to the cold water side of the front). The zusatzwind derived from the thermodynamic equation (Figure 18) is considerably weaker and the patterns differ. Which of the simplified computations gives the more realistic picture is difficult to decide at this stage. The answer to this question will be obtained from actual observations and from the computations with fine mesh window type (zoomed)

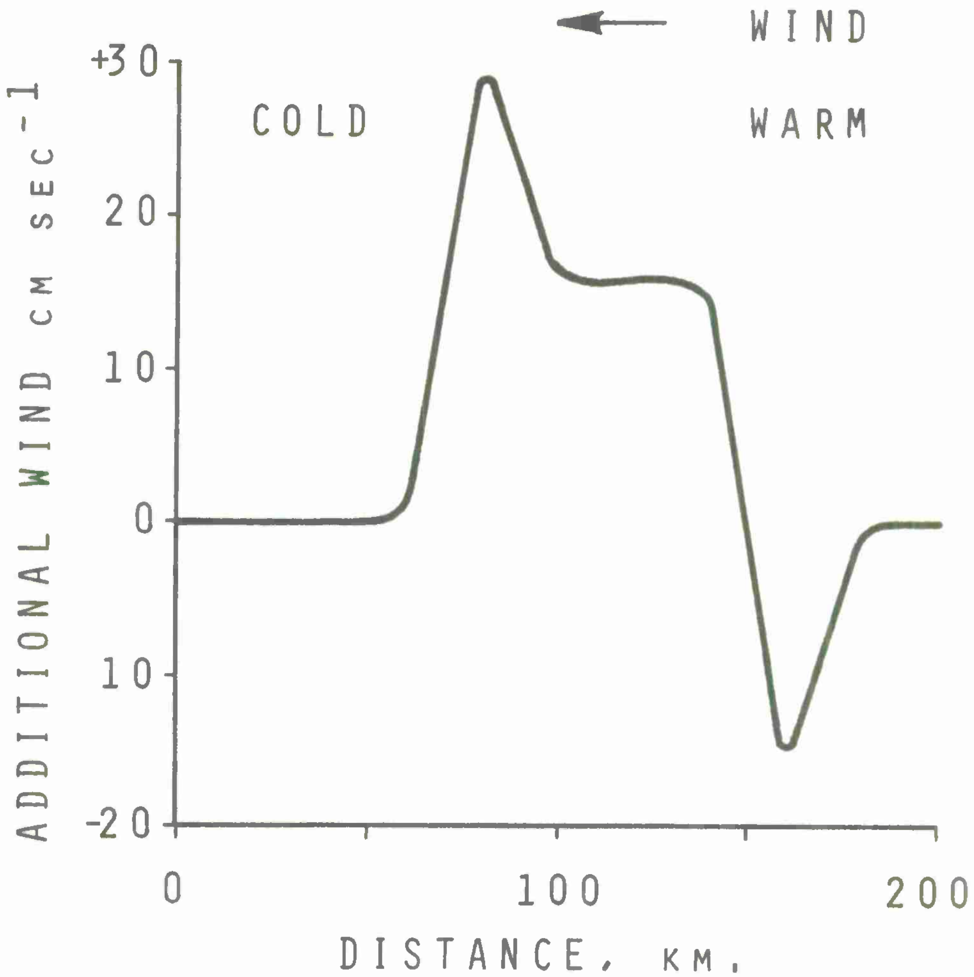

Figure 17. Zusatzwind From Equation of State, Wind Blowing From Warm to Cold Side.

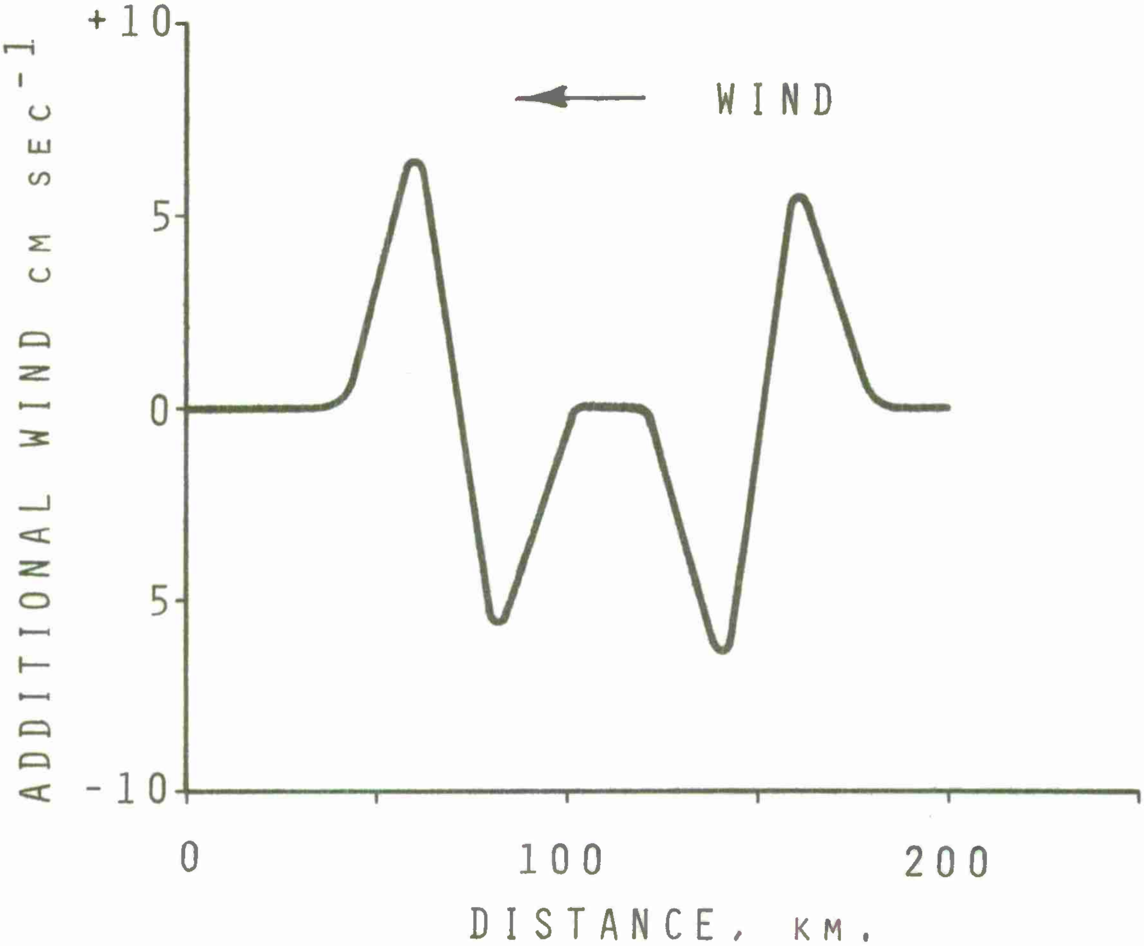

Figure 18. Zusatzwind From Thermodynamic Equation; Wind Blowing From Warm to Cold Side.

models. Attention could be called to the review of this subject by Roll (1965) who pointed out the occurrence of unusual mesoscale winds at sea which do not follow the classical relationship between reported surface wind and the computed gradient wind.

The low resolution of the computational grid in hemispheric atmospheric models inhibits the presentation of mesoscale features such as oceanic frontal effects and coastal effects. However, the sum of mesoscale features affects the hemispheric scale feedback and proper means must be found to include this.

There are clear indications in published data that the wind fields respond to oceanic fronts. One of the studies by Verploegh (1954) is reproduced in Figure 19 which shows a remarkable relationship between air/sea temperature and wind speed. Surface winds in excess of twice the gradient wind along the North Wall of the Gulf Stream (especially during cold air outbreaks) are also pointed out by Cormier and Kindle (1971).

The major oceanic fronts are usually much more pronounced than the ones assumed in the numerical example. Considering the effects of the oceanic fronts on the surface air (as demonstrated in this paper, and by empirical evidence not fully described here) it becomes apparent that the oceanic fronts are a significant feature in the placement of the mean

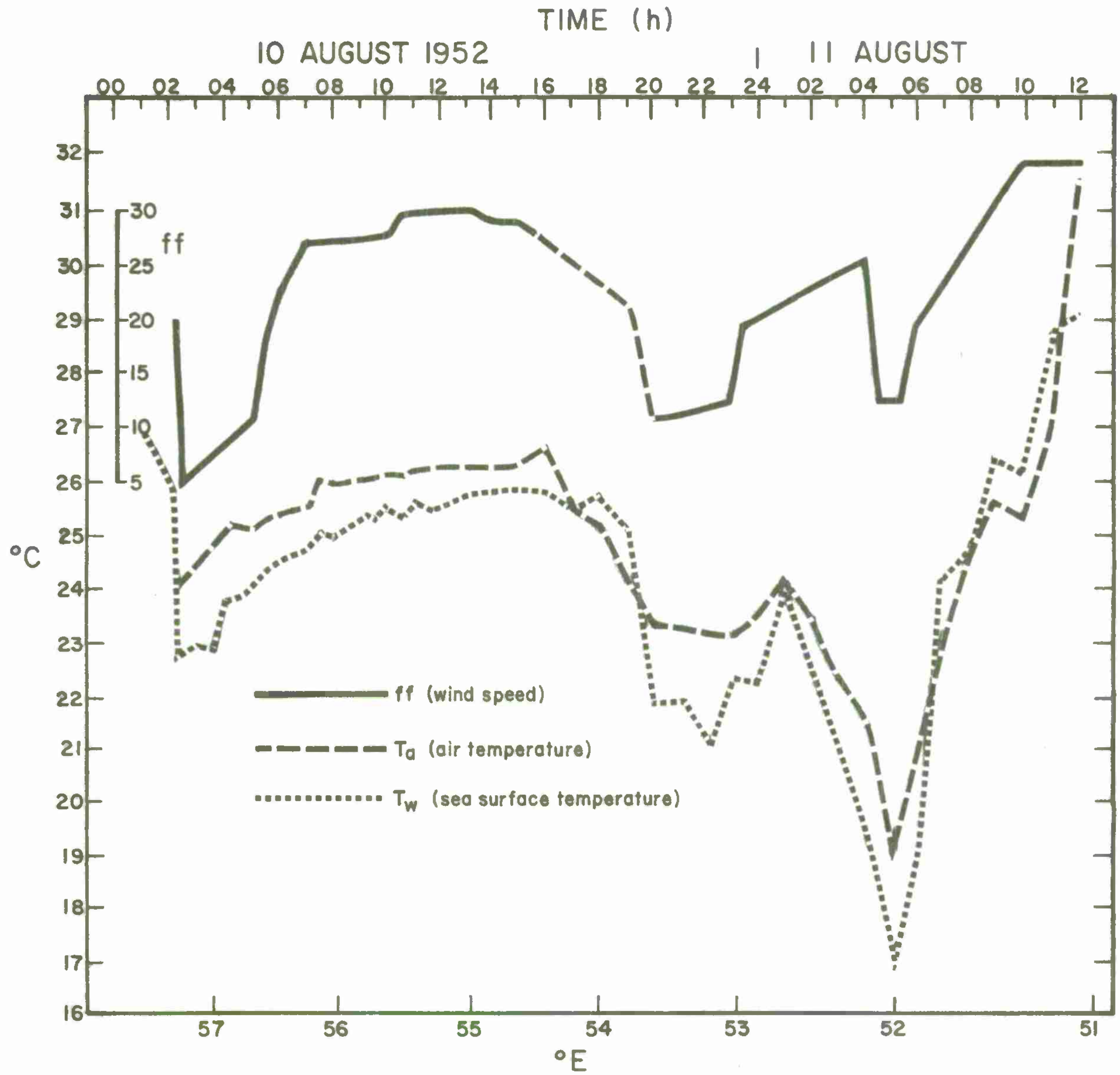

Figure 19. Changes of Meteorological Elements Along the Track of M/S "Borneo" Off Somali Coast, August 1952.

positions of atmospheric fronts near the vicinity of the former.

The theories and results of numerical computations should be verified by measurements in nature (e.g. frequent measurements on ships of opportunity while crossing oceanic fronts). The verification of the results is important as the same phenomena occurs, to a lesser degree, in all parts of the oceans when winds blow across sea surface isotherms. As the changes are most pronounced at oceanic frontal regions, more information is available in these pronounced thermal contrast areas.

6. CONCLUSIONS

a. The Åmot - Mosby theory of the response of surface air temperature (T_a) and water vapor pressure (e_a) to the corresponding properties of the sea surface is adapted for synoptic analysis and prediction of T_a and e_a over the oceans.

b. When this method is applied numerically to oceanic frontal region conditions, relatively rapid modification of surface air properties through heat and moisture exchange results. Several peculiarities of this modification process are observed in relation to winds and oceanic fronts (e.g. the "overshooting" of sea-air temperature difference and others). The peculiarities are physically explainable, however.

c. When the rate of change of sea surface temperature along a trajectory of the air remains constant, the sea-air temperature difference reaches an "equilibrium value" in about five hours.

d. The existence of an "equilibrium" difference between the surface air temperature and sea surface temperature (or respective water vapor pressures) does not mean a no heat and moisture exchange condition, but merely indicates that the supply of heat (and/or moisture) from the sea equals its removal (by turbulence) from the surface layers of the air.

e. Due to pronounced differential heating/cooling of the lower layers of the air in oceanic frontal regions, relatively sharp mesoscale pressure gradients are created which will modify the local winds.

f. The mean positions of atmospheric polar and arctic fronts over the oceans are located near the corresponding positions of the oceanic fronts. This study demonstrates that the oceanic fronts are significant controlling factors in the relative mean positions of atmospheric fronts.

g. Due to the sharp horizontal thermal gradients and pronounced effects of these gradients on the lower atmosphere, the oceanic frontal regions are the most ideal areas for the study and modelling of the mesoscale feedback from the ocean to the atmosphere.

h. As the mesoscale processes affect the large-scale features, proper methods must be devised for their inclusion into the large-scale numerical predictions of the atmosphere.

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